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Age of detrital zircon and titanite in the Meguma Group, southern Nova Scotia, Canada: Clues to the origin of the Meguma Terrane

T.E. KROGH [1] and J.D. KEPPIE [2]

[1] *Royal Ontario Museum, 100 Queen's Park, Toronto, Ont. M5S 2C6 (Canada)*
[2] *Department of Mines and Energy, P.O. Box 1087, Halifax, N.S. B3J 2X1 (Canada)*

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Abstract

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Precise U–Pb isotopic analysis of seventeen selected high-quality laboratory abraded zircon single grains, seven multigrain fractions of similar material and four paramagnetic titanite fractions record the ages of the various detrital grain types from two turbidite samples collected from the top and about 5 km below the top of the Goldenville Formation of the Meguma Group. The results show that the source of both sediments contained ~ 600 Ma and 2000 Ma old volcanic–plutonic components, whereas a third, 3000 Ma old component occurs exclusively in the upper sample and the youngest zircons and titanite are found in the lower sample. Lead–lead ages of bulk fractions at 2300 Ma and two nearly concordant single grains at 2040 Ma and 2140 Ma (upper sample), together with 2157 Ma (bulk fraction) and 2060 Ma and 2080 Ma single grain fractions (lower sample), indicate that the ~ 2000 Ma component dominates the pink–purple, euhedral grain types. Detrital titanite gives ages of 605 Ma (upper sample) and 552 Ma (lower sample), with the latter sample containing a minor ~ 1975 Ma component. Colourless zircons record this same age difference. The lower sample contained the youngest, concordant, single zircon grain (566 Ma) and 580–605 Ma bulk grain fraction, even though two concordant single grains have ages of 632 and 643 Ma. Estimated bulk fraction ages for certain zircon grain types from the upper sample fall in the range of 605 to 675 Ma. These data suggest that while the youngest rocks in the source region were removed by erosion during the early stages of deposition of the Goldenville Formation, the oldest source rocks were not unroofed until later in the depositional history. The 566 ± 8 Ma age of the youngest concordant zircon from the stratigraphically lower sample provides a maximum age for the deposition of this grain. With the base of the Cambrian placed between 545 and 530 Ma, the Goldenville Formation may straddle the Precambrian–Cambrian boundary. Graptolites from the immediately overlying Halifax Formation are of Early Ordovician (Tremadocian) age. The lower sample contains spherical zircons that may have been derived from mafic granulite with an estimated age of 2000–2200 Ma. With this one possible exception, all of the zircons analyzed appear to be igneous in origin and all are relatively euhedral, indicating that rocks of the ages referred to above probably outcrop in the source area. Anomalous discordance patterns imply that part of the 2960 Ma area was metamorphosed by the 2000 Ma event, and that both the 2960 and 2000 Ma regions were metamorphosed during the 600 Ma event. The complete lack of detritus with ages between 680 and 2040 Ma virtually eliminates a northern North American or Baltic Shield provenance. However, the available data are not sufficient to distinguish between a West African or Guyana Shield source.

Introduction

Juxtaposed against the Avalon Terrane, the meguma Terrane is the farthest outboard terrane in the northern Appalachians (Fig. 1) (Keppie, 1985, 1988). It is characterized by thick (?)Cambro-Ordovician turbidites of the Meguma Group (Schenk, 1970) overlain by shallow-marine

and continental rocks. The Meguma Group is made up of two formations: Goldenville and Halifax. The > 5500 m thick Goldenville Formation is mainly made up of metawackes with thin slate interbeds. It is overlain by the 500–4000 m thick Halifax Formation, which predominantly consists of slate with subordinate siltstone and wacke. The basal Halifax Formation contains Tremadocian grapolites, but no fossils with an unique age assignment have been recovered from the Goldenville Formation. In contrast, the Lower Palaeozoic rocks of the neighbouring Avalon Terrane are typically platformal in nature, containing a *Paradoxides* Cambrian trilobite fauna (Hutchinson, 1956; Palmer, 1973) and a *Selenopeltis*/trinucleid Ordovician trilobite fauna (Whittington, 1973).

Docking of the Meguma and Avalon Terranes is recorded by regional folding, cleavage development and concomitant low-grade metamorphism, which has yielded $^{40}Ar/^{39}Ar$ whole-rock phyllite plateau ages ranging from ~ 400 to 377 Ma (Keppie and Dallmeyer, 1987). Subsequent accretionary telescoping between the Meguma and Avalon Terranes was recorded by dextral shear zones with $^{40}Ar/^{39}Ar$ mica plateau ages ranging from ~ 375 to 287 Ma (Dallmeyer and Keppie, 1987). This was synchronous with granitoid plutonism ranging in age from ~ 375 to 270 Ma (Clarke and Halliday, 1980; Keppie et al., 1985; Cormier et al., 1987), and the deposition of Carboniferous strata as an overstep sequence across the Meguma–Avalon Terrane boundary; these strata were also involved in the Late Palaeozoic deformation. The proximity of the Meguma and Avalon Terranes in Late Silurian/Early Devonian times is signalled by the appearance of Rhenish–Bohemian brachiopod

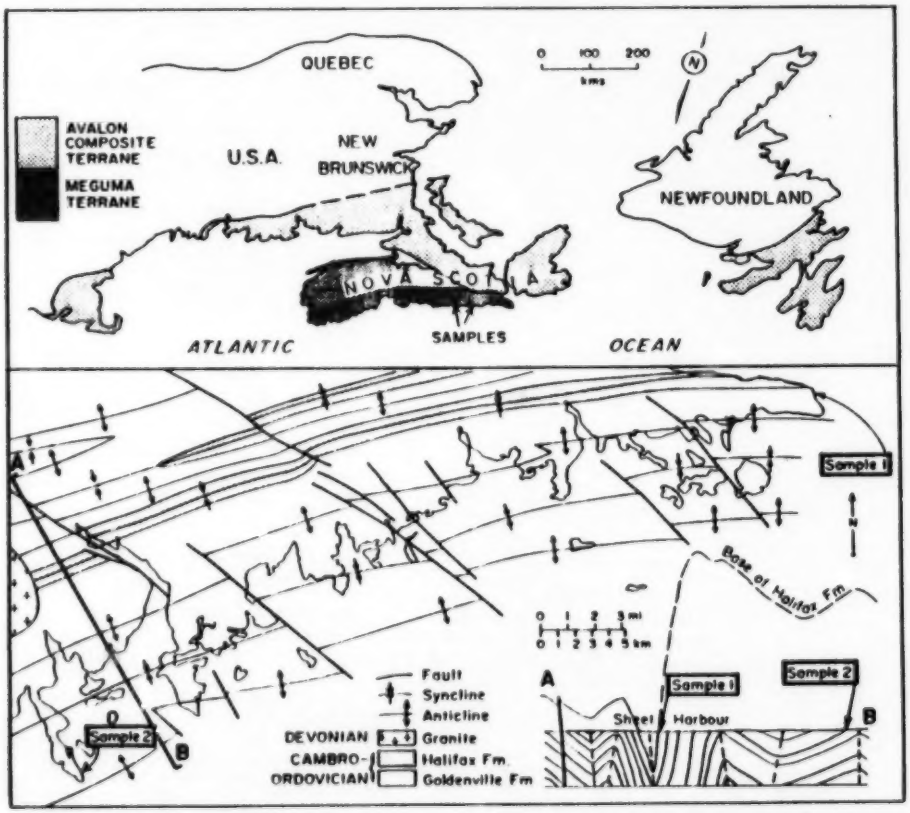

Fig. 1. Geological map showing the sample locations. The location of the Meguma Terrane is also shown.

fauna on both terranes in Pridolian rocks, this fauna spreading northwards into the Gaspe Peninsula by Emsian times where it mixed with the Eastern Americas Realm taxa (Boucot, 1975). The closure of the ocean between Euramerica and Gondwana by Late Devonian times was followed by the dispersal of freshwater vertebrates across the entire region (Young, 1981).

Early palaeozoic palinspastic reconstructions involving the position of the Meguma Terrane are based on lithological comparisons because palaeomagnetic and faunal province data are lacking for the Meguma Terrane. This has lead to several hypotheses for the origin of this Terrane:

(1) An intracratonic basin within the Avalon Zone (e.g. Dewey, 1969)

(2) A continental rise prism off northwestern Africa (e.g. Schenk, 1971, 1981, 1983)

(3) An intramontane basin (intradeep) within or marginal to northwestern South/Central American (e.g. Keppie, 1977, 1982).

Faunal provinces may provide some constraints on the palinspastic reconstructions during the Pridoli–Early Devonian. Thus, the Rhenish–Bohemian brachiopods which have been recorded in the Gaspe Peninsula, the Maritime Avalon and Meguma Terranes have also been found in Turkey Creek in Oklahoma, Cuidad Victoria in Mexico, in Western and Central Europe, and in northwestern Africa. In addition, Rhenish–Bohemian elements are also present in northern South America (Boucot, 1975). The impoverished nature of this fauna lead Boucot (1975) to suggest that it was distributed around a small ocean basin with restricted circulation and high evaporation rates. However, that North and South America were closer together than North America and Africa is suggested by the presence of the Eastern Americas Realm Early Devonian brachiopods in both North America and northern South America, but not in northern Africa (Boucot, 1975). Palaeomagnetic data bearing upon the Siluro-Devonian collision have suggested that Laurentia collided with South America (Miller and Kent, 1988) or Africa (Van der Voo, 1988). The relative position of the Meguma Terrane with respect to these major cratons during the Cambro-Ordovician has no such faunal or palaeomagnetic constraints. In addition, the Ordovician glaciation invoked in the Meguma Terrane by Schenk (1981) is negated by the reinterpretation of the "glacial deposits" as debris flows and mylonites (Keppie, 1982).

An approach to providing data bearing on this problem of the origin of the Meguma Terrane involves dating detrital minerals in the Meguma Group. Schenk (1970), based on detrital mineralogy which includes zircon and cassiterite, concluded that the Meguma Group was derived from a metasedimentary–metaigneous source. Detrital muscovite dated by the K–Ar technique yielded ages of 485–506 ± 20 Ma (Poole, 1971); however, it is clear that this does not record the age of the source, but is closer to the age of deposition. Clarke and Halliday (1985) have dated metawacke and metapelite from the Meguma Group using the Sm/Nd technique and produced a mean crustal residence age of 1773 ± 95 Ma. However, this technique would merely produce a mixed age if the source was made up of several ages of rocks, which may themselves have originated from several sources. In order to obtain precise age information about the provenance of the Goldenville Formation, we have analyzed seventeen individual zircon grains, seven homogeneous groups of zircon grains and four multigrain fractions of titanite from two samples (Fig. 1). Grains were selected to represent the full range of colour and morphology visible in the zircon populations extracted from two 50 kg samples collected at the Halifax–Goldenville contact and about 5 km below it (samples 1 and 2 respectively). The results show that the source was indeed polygenetic, and they also provide a maximum deposition age for most of the formation.

Analytical techniques

Each sample was crushed using a jaw crusher and disc mill and minerals were separated using a Wilfley Table, heavy liquids and a Frantz Isodynamic Separator. All crushing and grinding plates were removed and cleaned with alcohol and an electric brush prior to use to avoid any possibility of even single grain contamination. Zircon concentrates were further separated into magnetic and size fractions but, as noted in Table 1, only the very least paramagnetic fractions were used in

TABLE 1

Isotopic data for zircon and titanite grains in Goldenville Formation sediments

ZIRCON FRACTIONS	WEIGHT [mg]	U [ppm]	COMMON Pb[a] [pg]	$^{207}Pb/^{206}Pb$	$^{206}Pb/^{204}Pb$	$^{208}Pb/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$ AGE (Ma)
GOLDENVILLE FORMATION Z-1									
1. NO, PINK, EUH	0.001	154	21	0.21753	298	0.1474	0.57961	17.3843	2963
2. NO, BROWN, OBLATE	0.030	171	9	0.20105	18883	0.1372	0.51767	14.3501	2835
3. NO, PINK, EUH + BLACK DOTS	0.002	189	15	0.21535	995	0.2387	0.54330	16.1320	2946
4. NO, PINK, EUH	0.005	120	8	0.21652	2508	0.3073	0.52262	15.6023	2955
*5. NO, PURPLE, EUH	0.040	251	69	0.14381	3550	0.1426	0.40049	7.9412	2274
6. NO, LIGHT BROWN, EUH	0.002	414	9	0.13292	2582	0.2063	0.38701	7.0926	2137
7. NO, PINK, EUH	0.002	328	32	0.12588	454	0.1266	0.36854	6.3968	2041
*8. NO, COLOURLESS, EUH	0.089	177	14	0.07585	8552	0.1517	0.12583	1.3160	1091
*9. NO, COLOURLESS, EUH, LONG	0.142	237	18	0.06474	11484	0.2181	0.10256	0.9155	766
*10. MIO, 1.0A, BROWN, TITANITE	0.186	132	500	0.06166	316	-	0.10074	0.8565	662
*11. MIO, 1.7A, BROWN, TITANITE	0.158	101	556	0.06086	190	-	0.09942	0.8343	634
GOLDENVILLE FORMATION Z-2									
*12. NO, PINK, ROUND	0.103	152	55	0.13912	6499	0.1227	0.37435	7.1808	2216
*13. NO, DARKER PINK ROUND	0.090	342	94	0.13805	7667	0.1349	0.38982	7.4200	2203
14. NO, LIGHT PINK, BLOCK, EUH	0.005	102	8	0.13357	1436	0.1511	0.35871	6.6062	2145
15. NO, LIGHT PINK, BLOCK, EUH	0.007	81	4	0.16053	3958	0.1541	0.45260	10.0180	2461
16. NO, LIGHT PINK, EUH	0.008	274	20	0.12718	2456	0.1168	0.37101	6.5057	2059
*17. NO, LIGHT PINK, EUH	0.140	228	105	0.13444	6957	0.1309	0.37992	7.0425	2157
18. NO, COLOURLESS, BLOCK, EUH	0.008	75	7	0.12254	1550	0.0910	0.29880	5.0486	1994
19. NO, COLOURLESS, OBLATE, SPHEROID	0.009	24	4	0.06038	278	-	0.08429	0.7018	617
*20. NO, COLOURLESS, EUH+ BUBBLES	0.100	188	44	0.06989	2807	0.1665	0.10811	1.0418	925
*21. MIO, 1.0A, TITANITE	0.565	124	2325	0.06567	196	-	0.09816	0.8888	796
*22. MIO, 1.7A, TITANITE	0.888	147	3226	0.07659	295	-	0.11398	1.2037	1110
23. NO, COLOURLESS, EQUANT	0.002	138	6	0.05898	374	0.2547	0.09075	0.7380	566
24. NO, COLOURLESS, EQUANT	0.010	43	6	0.12700	1522	0.1557	0.32031	5.6088	2057
25. NO, COLOURLESS, EQUANT	0.004	177	4	0.06109	1275	0.1898	0.10353	0.8721	643
26. NO, TRACE BROWN, EQUANT	0.008	48	4	0.12886	2324	0.1956	0.37504	6.6636	2083
27. NO, COLOURLESS, ROUNDED ELONGATE	0.007	101	6	0.06216	763	0.1933	0.09216	0.7898	680
28. NO, TRACE BROWN, ELONGATE	0.002	362	7	0.06081	739	0.1913	0.09970	0.8359	632

Notes to Table 1

[a] Samples containing a selected group of similar grains, other data are for single grains. N = non-magnetic, M = magnetic number following this abbreviation refers to degree of tilt on the Frantz Isodynamic Separator at full current). For titanite, Frantz magnetic current = 1.0 A or 1.7 A.
EUH = euhedral.
All grains are abraded to remove surface-correlated lead loss.

[b] Total common lead present; includes components from sample and blank.

[c] Corrected for fractionation, blank and common lead in the spike plus sample. The isotopic composition of common lead in the sample is calculated using the Stacey and Kramers model (1975). Single grains were corrected for 4–6 pg of blank lead. Multigrain samples were corrected for 12 pg of blank lead. In correcting for fractionation, $^{207}Pb/^{206}Pb$ and $^{238}U/^{235}U$ are increased by 0.13%/amu.

[d] Measured values corrected for spike contribution.
^{238}U decay constant = 0.15513×10^{-9}/yr
^{235}U decay constant = 0.98485×10^{-9}/yr
$^{238}U/^{235}U = 137.88$

the final selection of the highest quality grains under a binocular microscope. All grains selected were then abraded prior to analysis (Krogh, 1982) to reduce or eliminate surface-correlated lead loss associated with geologically leached surfaces. Zircons were dissolved and analyzed as described in Krogh (1973), with the exception that smaller dissolution capsules and columns for the resin were used in some cases (i.e. capsules of 0.5 cm^3 vs. > 5 cm^3 and columns of 0.05 cm^3 vs 0.5 cm^3).

Total procedure Pb blanks of 6–12 picograms (pg) and 2–6 pg were measured during the early multigrain analysis and the more recent single grain analysis respectively. A mixed ^{205}Pb–^{235}U isotopic tracer was added prior to a 5 day digestion at 220°C. The initial procedure whereby Pb and U were analyzed sequentially on the same filament using first silica gel and then Ta_2O_5 was later replaced, both elements subsequently being loaded together on silica gel. Lead was then measured at successively higher temperatures between 1450° and 1600°C, and uranium between 1600° and 1800°C. Ratios of ^{207}Pb to ^{204}Pb were measured using a Daly detector in an integrating mode, as were all ratios for smaller samples. In general it was possible to measure the more critical $^{207}Pb/^{206}Pb$ ratio on the Faraday detector if more than 12 picograms of ^{207}Pb were present.

Analytical uncertainties due to measurement errors alone are less than 1.0 and 0.2% (2 σ) for the Pb/U and $^{207}Pb/^{206}Pb$ ratios respectively, corresponding to about ±4 Ma in the $^{207}Pb/^{206}Pb$ age. The size of the error ellipses for most samples in Figs. 2–5 reflects this uncertainty. However, in cases where the sample to background ratio is low, our error estimate has been enlarged to reflect the fact that the isotopic composition of the common lead is not precisely known and the precision on the measurement of the amount of background is more critical. Thus, the errors in analysis 1 (Fig. 3) and analysis 7 (Fig. 4) have been increased by a factor of 2 and 1.5 respectively. The errors in all of the analysis in Fig. 5 for the younger, often low-uranium colourless grains selected to give a high probability of concordance were doubled even though most contain between 4 and 6 pg of common lead (column 3, Table 1), and the isotopic composition of this lead is well known because it is essentially all blank. It should be noted that whereas ratios and ages have the accuracies quoted, due to uncertainties in sample weight uncertainties in concentrations may be in error by as much as 20%.

A test of the overall reproducibility is included here as this is the only manner of truly demonstrating uncertainties when sample sizes are reduced to unprecedented low levels. For this test we chose to analyze six single zircons from a plagiogranite in a gabbro of ophiolitic affinity (Wandsworth Gabbro, Burin Peninsula, Newfoundland). The grains are relatively young (763 ± 2 Ma) and therefore more difficult to analyze than most. Natural geological dispersion, due to inheritance, is unlikely due to the source. Previous results are given in Krogh et al. (1988). The results shown in Fig. 2 demonstrate that there is no

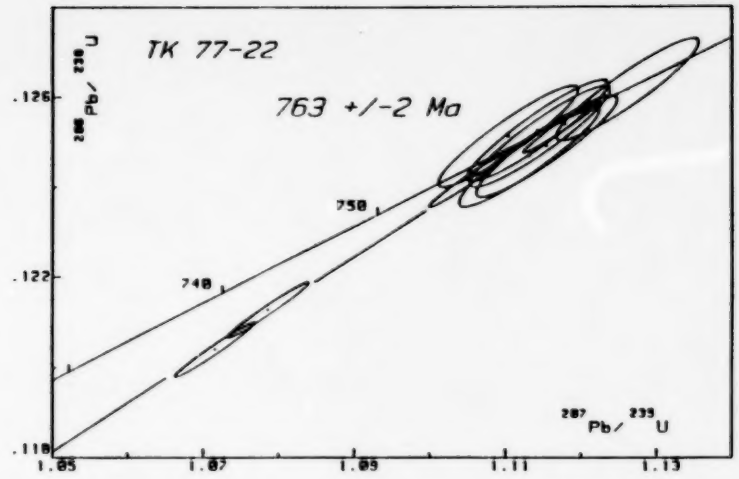

Fig. 2. $^{206}Pb/^{238}U$–$^{207}Pb/^{235}U$ plot of six single zircon analyses (dots with the larger error ellipses) compared with multigrain analyses (dots with the smaller error ellipses) previously reported in Krogh et al. (1988) from the Wandsworth Gabbro, Burin Peninsula, Newfoundland.

systematic bias for small samples; the mean $^{207}Pb/^{206}Pb$ age is 762.4, whereas larger fractions give an upper intercept at 763 ± 2 Ma. The $^{206}Pb/^{238}U$ age has a mean of 760 Ma and a range of 9 Ma, or 1.2% (four of the six fall between 759 and 761 Ma inclusive). If the error on the original line is increased from ± 2 to ± 4 Ma, five of the six single grains fall within this limit. The sixth analysis that plots above the curve to the left of the main data set may illustrate a minor problem yet to be eliminated in small sample analysis. A small amount of organic material usually percepti-

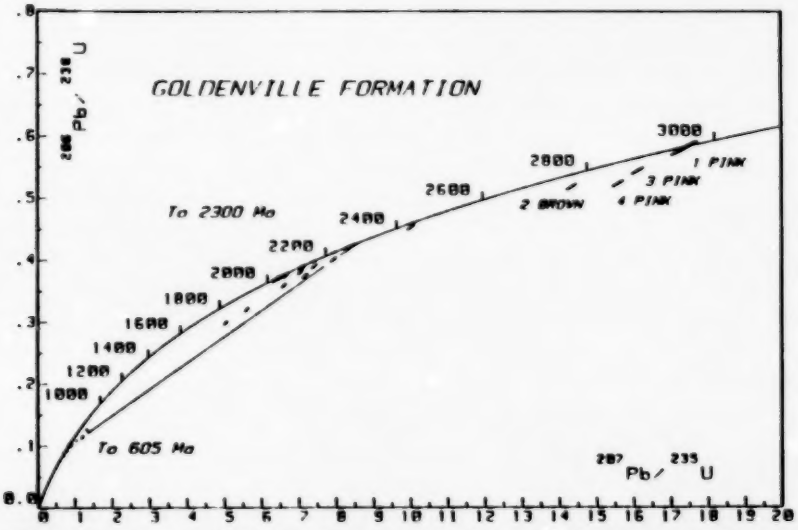

Fig. 3. $^{206}Pb/^{238}U$–$^{207}Pb/^{235}U$ plot of the Goldenville Formation detrital zircons and titanite. In this figure and in Figs. 4 and 5, data for sample 1 are numbered 1–11 and for sample 2 are numbered 12–28. In addition, data for multigrain fractions are denoted by circles drawn over the error ellipse and data for grains which are highly rounded due to growth habit or sedimentary abrasion, or both, are denoted by the letter *R*.

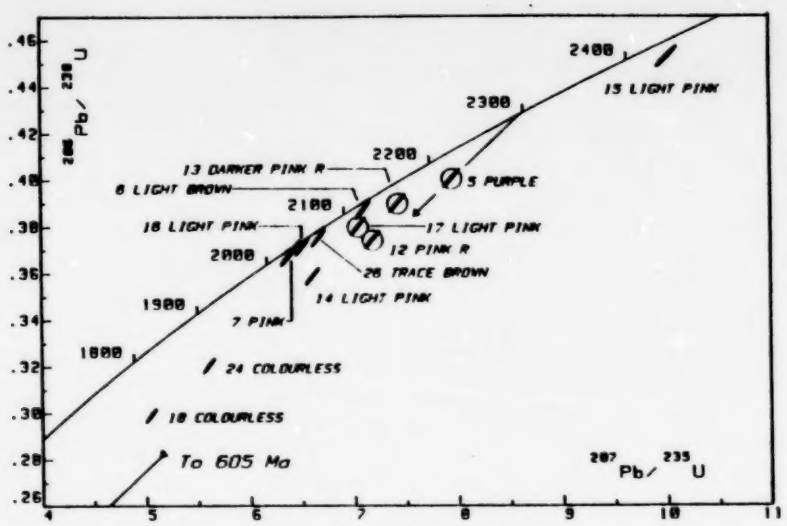

Fig. 4. Enlargement of part of Fig. 3. Symbols are the same as in Fig. 3.

ble early in the run may persist and make a significant contribution at the small ^{204}Pb mass position. This causes a spurious overcorrection for common lead and displaces a concordant point to the left and above the Concordia. This problem is generally avoided if the load is well centred on the filament and the $^{207}Pb/^{204}Pb$ ratio is measured at progressively higher temperatures and only the latest (and highest) ratios are used. This test also confirms the current low and consistent analytical Pb blank because the total common lead observed only varied between 3.0 and 4,1 pg. It also demonstrates that a combination of selection and abrasion can consistently yield near-concordant results.

Results

Uranium–lead isotopic data for the single and multigrain fractions are given in Table 1 and shown in Figs. 3–5.

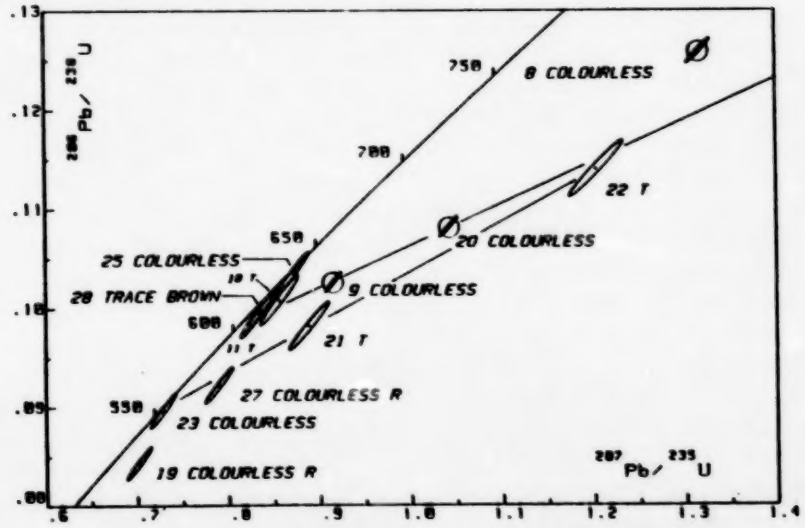

Fig. 5. Enlargement of part of Fig. 3. Symbols are the same as in Fig. 3.

Sample 1, from the top of the Goldenville Formation

A combination of single grain and multigrain analysis of a single grain type is useful in determining both the individual source ages and the relative contribution of each source. A multigrain fraction consisting of about twenty simple euhedral purple–pink prisms with a length to width ratio of about 2 to 1 from sample 1 appeared to represent a single population which might be from a single source region (analysis, 5, Table 1, Fig. 4). However, the single grains analyzed from this group indicate that sources with ages of both 2040 and 2963 Ma are present (analyses 7 and 1, Table 1, Figs. 3 and 4). Because the multigrain fraction (analysis 5, Fig. 4) plots near 2300 Ma it is certain that the younger source is the major contributor of this grain type. The analysis of a brown euhedral grain revealed that the younger terrane also contains 2140 Ma old rocks (analysis 6, Fig. 4). A similar grain (analysis 2, Fig. 3) has an apparent Pb/Pb age of 2835 Ma, although it could have been derived from a 2960 Ma old outcrop that suffered lead loss during the ~ 2000 Ma event. The relatively high degree of discordance exhibited by the data for two other low-uranium, euhedral pink grains (analyses 3 and 4, Fig. 3) suggests that zircons in parts of the ~ 2960 Ma source may have suffered lead loss during the ~ 600 Ma event identified below.

It is generally observed that geologically young zircons (say less than 0.5 Ga) are low in colour as if much of the colour present in older zircons results from accumulated radiation damage. This concept is supported by the fact that in a single population the low-uranium grains have the lowest colour and by the observation, that xenocrystic grains annealed in younger granite or in the laboratory are generally colourless, regardless of age. Thus we chose to analyze two multigrain, colourless fractions. Data for short and long euhedral sharp-faceted crystals are shown as analyses 8 and 9 respectively in Fig. 5 and Table 1. Seen in perspective in Fig. 3 it is certain that most grains of this type are relatively young. Assuming that the most probable admixed older component is represented by the multigrain fraction (analysis 5), we can project from 2300 Ma to obtain a probable average age for the young component of ~ 605 Ma for the elongate grain types (analysis 9) and ~ 675 Ma for the more equant euhedral types (analysis 8). This difference could be real if the two grain types represent plutonic and volcanic igneous sources respectively. If the data point for the short grains (analysis 8) is placed on a chord through 2050 Ma (the lowest age in the ~ 2000 Ma group), the lower intercept yields an average source age of ~ 647 Ma. Thus, the average age for the short type of zircon grains probably lies between ~ 647 and 675 Ma.

Titanite from sample 1 has a round, frosted appearance, as if this much softer mineral has been abraded during sedimentary transport. Data for two multigrain paramagnetic fractions (analysis 10 T and 11 T, Table 1, Fig. 5) plot very close to the Concordia adjacent to data for acicular zircons (analysis 9, Fig. 5), as if only one source age is present. In fact, these data and those for the multigrain fraction (analysis 5) are colinear between ~ 605 and 2300 Ma and this line also includes data for a particular grain type from sample 2 (described below). For reference, segments of this line are shown in Figs. 3–5.

Sample 2, collected about 5 km below the boundary between the Halifax and Goldenville Formations

Sample 2 also contains a coherent population of pink–purple simple prisms with length to width ratios of about 2 to 1. The multigrain fraction from this grain type (analysis 17, Table 1, Fig. 4) indicates a Pb–Pb age of 2157 Ma, but the single grains analyzed again demonstrate that more than one age is present. As seen in Fig. 4 and Table 1, analyses 16 and 26 are almost concordant and indicate that rocks with ages close to 2059 and 2083 Ma occur in the source region, whereas analysis 15 is more discordant and has a Pb–Pb age of ~ 2460 Ma, or ~ 2600 Ma if placed on a chord through 2070 Ma. Again, the younger component is the dominant member in the mixture because the data for the multigrain fraction plots near that for the youngest grains present.

Two multigrain fractions of rounded grains that differed only in their degree of pink colouration

were analyzed to give data points 12 and 13 (Table 1, Fig. 4). Although these rounded grains could have been produced by extreme abrasion of igneous zircons during sedimentary transport, they also strongly resemble spherical euhedral metamorphic crystals that occur in mafic granulite facies rocks. If so, parts of this source region may have undergone a granulite-facies metamorphic event about 2200 Ma ago, or alternatively this event could have occurred about 2000 Ma ago if some similar but much older grains were included in the multigrain sample. Tests on single grains of this population were not carried out due to the scarcity of this grain type.

Analysis 18 and 24 (Table 1, Figs. 3 and 4) represent the only colourless grains analyzed that have Pb–Pb ages as old as 2.0 Ga. These data are peculiar in their high degree of discordance because each grain was selected as a high-quality non-magnetic, crack-free grain and because each has a relatively low uranium content (75 and 43 ppm respectively). These observations strongly suggest that part of the ~ 2000 Ma region underwent a metamorphism about 600 Ma ago and that this event annealed out the radiation-induced colour and produced the high level of discordance observed. Analysis 14 is also anomalously discordant for a high-quality low-uranium grain (102 ppm, Table 1, Fig. 4). In this case, the light pink colour must have been either retained or developed after the metamorphic event.

Only one multigrain fraction of colourless zircon was analyzed from sample 2 (analysis 20, Table 1, Fig. 5). As previously described, the data representing this coherent group of euhedral crystals with bubble-like inclusions plot on the line between 605 and 2300 Ma, as if most, but not all grains of this type are about 600 Ma old. The youngest, concordant, colourless zircon has a Pb–Pb age of 566 ± 8 Ma (analysis 23, Table 1, Fig. 5). Three, single colourless zircon grains gave almost concordant data with Pb–Pb ages of 566, 632 and 643 Ma (analysis 23, 28 and 25, Table 1, Fig. 5). Data for two other colourless zircon grains (analyses 19 and 27, Table 1, Fig. 5) gave less concordant and hence less precise age information. They may have an unusually high degree of recent lead loss with primary ages of about 620 and 680 Ma or they may have originated from younger intrusions of, say, 550 and 500 Ma old and contain a trace of older core component.

Two multigrain fractions of frosted titanite rounded during sedimentary transport (analyses 21 T and 22 T, Table 1, Fig. 5) indicate that at least two ages are present. Combined with the zircon data described above, a two-component mixing line with ages of 552 ± 5 and 1975 ± 50 Ma is defined. The titanite data could indicate 2000 Ma old grains strongly but not completely reset during a 550 Ma event or may represent a mixture with most of the grains originating from the younger source. Single grain analyses to resolve this, though technically feasible, were not considered warranted at this time. Although less certain, the 552 ± 5 Ma age of the younger titanite provide a lower limit for the time of deposition of the lower sample.

Interpretation

While ~ 600 Ma old zircon and titanite and 2000 Ma old zircon are present in both Goldenville samples, the youngest concordant zircon (566 ± 8 Ma) and titanite (552 ± 5 Ma, lower intercept) are only present in the stratigraphically lowest sample, and the oldest zircons (analyses 1–4, with the oldest $^{207}Pb/^{206}Pb$ age of 2963 Ma) are only present in the stratigraphically highest sample. These data suggest that while the youngest rocks in the source region were removed by erosion during the early stages of deposition of the Goldenville Formation, the oldest source rocks were not unroofed until later in the deposition of the Goldenville Formation.

The youngest ages provide a maximum age for an horizon about 500 m above the lowest exposed unit of the Goldenville Formation. The age of the Proterozoic–Cambrian boundary is at present rather uncertain. Harland et al. (1982) estimated it to be ~ 570 Ma (no errors suggested), while Odin (1982, 1986) favoured 530 ± 10 Ma. In the Avalon Composite Terrane of Newfoundland, a 565 ± 3 Ma U–Pb zircon age has been recorded from a tuff in the lower part of the Mistaken Point Formation (G. Dunning, pers. common., in Benus, 1988). This formation is overlain by two late Proterozoic units (St. John and Signal Hill Groups,

~ 1.6 km and 3–8 km thick), which are in turn overlain by Lower Cambrian rocks (King, 1980). In Morocco, a rhyolite in the late Proterozoic Ouarzazate Group has yielded a U–Pb zircon age of 578 ± 15 Ma (Juéry et al., 1974, 1975). This unit is overlain by the late Proterozoic, "série des calcaires inférieurs" which contains interstratified volcanic rocks associated with the Jbel Boho syenite dated at 534 ± 10 Ma (U–Pb zircon age) (Ducrot and Lancelot, 1977). The overlying "série lie-de-vin", containing late Proterozoic (Vendian) stromatolites (Charlot, 1978; Schmitt, 1978) is, in turn, overlain by the fossiliferous Lower Cambrian (lower Atdabanian) "série des calcaire supérieurs" (Sdzuy, 1978; Debrenne and Debrenne, 1978). In southern Britain, a granophyre dated at 531 ± 3 Ma (U–Pb zircon age) (Compston et al., in Cowie and Johnson, 1985) is unconformably overlain by Lower Cambrian sedimentary rocks containing fossils as old as latest Tommotian or earliest Atdabanian (Cope and Gibbons, 1987; this granophyre is therefore either Tommotian or late Proterozoic in age. Thus, the base of the Cambrian can be no older than ~ 545 Ma, and may be as young as ~ 530 Ma. The significance of this is that although most of the Goldenville Formation is Cambrian, its lower part may straddle the Proterozoic–Cambrian boundary.

The presence of sharp-faceted, unabraded needle-like zircons in the younger fractions suggests that the source of these grains is probably nearby. This is supported by the exclusively younger ages of detrital titanite, a mineral which is more prone to abrasion during sedimentary transport than zircon. Additionally, titanite is scarce to absent in granulite-facies metamorphic regions. The euhedral shape of the young zircons is most consistent with a plutonic/volcanic igneous source. An igneous, or igneous–metamorphic source for the titanite is suggested by the coincidence of its age with that of the needle-like zircon. The age of the rocks in the younger source region was between ~ 550 and 600 Ma during deposition of the lower Goldenville sample, but had shifted to between ~ 600 and675 Ma by the end of Goldenville deposition.

The sharp-faceted nature of most of the ~ 2–3 Ga zircon grains probably indicates that both the younger and older sources are exposed in the source region. The predominance of the younger source is indicated by the absence of the oldest component in one sample and a Pb–Pb age closer to 2 Ga in both multigrain samples. The morphology of these 2–3 Ga zircons would be most commonly found in a rhyolite or ignimbrite, although virtually any other igneous source could be involved. The stratigraphically lower sample also contains pink, rounded grains of a type commonly found in mafic granulite-facies rocks. An age of about 2200 Ma for multigrain fractions may date this metamorphic event or alternatively it could be as young as 2000 Ma if a small amount of older material was included in the sample. That a 2 Ga metamorphic event may have been superimposed upon the 3 Ga rocks is indicated by analysis 2, which falls on a chord between 2 and 3 Ga (Fig. 3). Similarly, the superimposition of a ~ 600 Ma metamorphic event on the 3 Ga rocks is suggested by the discordant position of analyses 3 and 4. The high degree of discordance observed in several low-uranium grains (analyses 18 and 24) strongly suggests that parts of the ~ 2 Ga terrain were subjected to metamorphism about 600 Ma ago. This is supported by the ~ 600 Ma old titanite grains which indicate some inheritance from the 2 Ga time period.

As this study progressed, the apparent lack of Grenvillian-aged zircons (~ 1 Ga) became increasingly evident as well as geologically important. To further test or verify this observation, we chose to analyze colourless grains to bias our selection toward younger ages and to select grains that had been rounded by abrasion to bias our sample toward the older side of this young population. No ~ 1 Ga age grains were found, nor were grains from the 750 Ma Pan-African units discovered.

Combining these data allows a reasonable picture of the source region to be constructed. This region consisted of 550–675 Ma old volcanic/plutonic/metamorphic rocks lying close to the Cambrian coast. These rocks may have been juxtaposed against an 2040 Ma old igneous/high-grade metamorphic basement (or any derived immature sediments) that may itself have been superimposed upon a 2960 Ma old igneous protolith.

Correlations

Based upon the detrital zircon data, the present position of the Meguma Terrane adjacent to the Avalon Composite Terrane would appear to be anomalous. Thus, while late Proterozoic igneous rocks yielding U–Pb ages between ~ 760 and 550 Ma are ubiquitous (Jamieson et al., 1986; Krogh et al., 1988; Keppie et al., 1989) and could be the source of the younger zircons and titanite in the Meguma Group, they overlie a metamorphic/igneous basement with ages unlike those in the Goldenville detrital zircon population. Thus, in southern New Brunswick, this basement has yielded U–Pb ages of ~ 814 Ma (concordant zircon from Brookfield Gneiss) which are inferred to be the age of metamorphism, and dubious, upper intercept ages of ~ 1640 Ma (discordant zircon from Brookfield Gneiss) which are inferred to represent the source age (Olszewski and Gaudette, 1982). In the Cobequid Highlands, basement granitoid gneisses of the Mount Thom Complex have given a 934 ± 82 Ma, 5-point, Rb–Sr whole-rock isochron age (no MSWD reported by Gaudette et al., 1984) which was interpreted by Gaudette et al. (1983) and Donohoe (1983) providing a minimum age for the high-grade metamorphic–deformational event. In northern Cape Breton Island, a syenite intrusive into gneisses of the Pollets Cove Brook Group (part of the Pleasant Bay Complex inferred to be part of the basement to Avalon Composite Terrane by Keppie et al. (1989)) has yielded discordant, U–Pb zircon results which correspond to a chord with lower intercept age of 918 + 78/ − 179 Ma and an upper intercept age of 1172 + 135/ − 73 Ma (Barr et al., 1987), here interpreted as the ages of metamorphism and intrusion, respectively. These ages from the basement are not represented in the detrital zircon population in the Meguma Group. Work in progress on Precambrian sediments in the Avalon Composite Terrane indicate a wide variety of ages that are not represented in the Meguma Group. Also, the 2–3 Ga source of the older zircons in the Meguma Group is either not present in the Avalon Terrane or is not presently exposed.

Detrital zircon and monazite in Cambro-Ordovician metasedimentary rocks from the southernmost French Massif Central (Montagne Noire) and from the Bohemian Massif have yielded the following ion microprobe (SHRIMP) U–Pb zircon ages: (1) inferred magnetic ages of 3.84, 3.15, 2.90, 2.76, 2.65, 2.60, 2.59, 2.54, 2.51, 2.15–1.75, 1.85, 1.00 and 0.75–0.53 Ga, and (2) inferred metamorphic ages of 2.59, 1.94, 1.00, 0.6, 0.46, 0.44 and 0.32 Ga (Gebauer et al., 1989). Other ion microprobe data for detrital zircons from high-grade Moldanubian metasediments have yielded discordant data with $^{207}Pb/^{206}Pb$ ages of between ~ 1750 and 2680 Ma (Kroner et al., 1988). Ion microprobe data from two different paragneisses from the southern Alps gave ages of 1.1, 1.9, 2.25, 2.6 and > 2.7 Ga (Grünenfelder et al., 1984). Thus, while the 3.0–2.0 Ga detrital zircons in the Cambro-Ordovician sediments in the European Hercynides and the Meguma Group are comparable, it is significant that the 1.0 Ga zircons are absent in the Meguma Group. Unfortunately, similar studies have not been conducted in other sedimentary units which might correlate with the Meguma Group.

Possible igneous/metamorphic source areas for the ~ 600 Ma zircon and titanite are ubiquitous throughout the eastern Appalachians, Europe, northwest Africa and northern South America, all of which have been affected by the Avalonian–Cadomian–Pan-African–Brasiliano Orogenies (Fig. 6). Further, while late Proterozoic volcanic rocks appear to be absent from the platformal cover on the Baltic craton, (?)late Proterozoic tholeiitic mafic dyke swarms are present in the nappes of the Middle Allochthon in central Scandinavia (Solyom et al., 1979). Thus, the ~ 600 Ma ages from the detrital zircons and sphene in the Meguma Group could have been derived from a wide geographical area along the eastern half of the orogen, although Baltica might be less favourable.

A source for the 2 and 3 Ga zircon in the Meguma Group may be found in northwestern Baltica, in the Icartien block in the Channel Islands and northwestern France, in the Aquitaine block in the Bay of Biscay, in the West African craton, in the Guyana craton and surrounding terranes such as Garzon, Sierra Nevada, Puqui, Cosinas, Perija, Santander, Quetame-Merida,

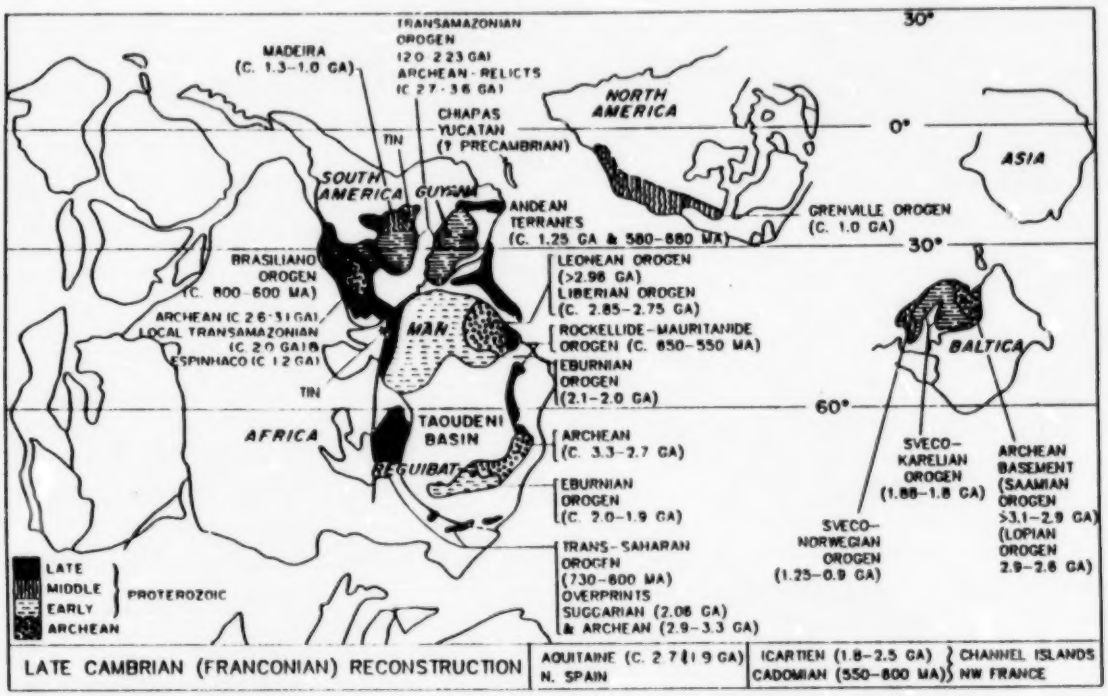

Fig. 6. Distribution of Proterozoic and Archaean rocks dated by U–Pb and Rb–Sr in and adjacent to the circum-Atlantic Palaeozoic orogens plotted on a Late Cambrian reconstruction modified after Scotese (1984). The Channel Islands, northwestern France and Aquitaine, and northwestern Spain all fall below the bottom edge of the figure.

Cajamarca and Payande, and in the Chiapas–Peten–Yucatan block of Central America (Fig. 6). The existing geochronological data for these areas will be discussed below.

The Baltic Shield consists of Archaean ortho- and paragneisses formed mainly during the Lopian (Rebolian) Orogeny and yielding zircon ages of between 2.6 and 2.9 Ga (Skiold, 1979, 1987; Gaal and Gorbatchev, 1987). Fragmentary evidence of an older Saamian Orogeny (> 3.1–2.9 Ga) is present. These Archaean rocks are overlain by sedimentary and volcanic rocks with U–Pb ages of > 2.2–1.86 Ga which were involved in the Svecokarelian Orogeny across central Scandinavia, dated between 1.88 and 1.8 Ga by U–Pb (Skiold, 1987). In southern Scandinavia, these rocks were reworked by the Sveconorwegian Orogeny between 1.25 and 0.9 Ga (Gaal and Gorbatchev, 1987).

The Icartien block in the Channel Islands and neighbouring northern France consists of granulitic gneisses which have yielded upper intercept U–Pb ages on zircons of 1790–2500 Ma, with individual data points plotting up to a third of the way down the chords towards 285–384 Ma (Auvray et al., 1980; Vidal et al., 1981). These gneisses were intruded by granitoid plutons dated at between 500 and 650 Ma (Jonin and Vidal, 1975) and were in turn unconformably overlain by the Lower Cambrian. These units are inferred to underlie the Armorican Massif.

Granulites recovered from the southern Bay of Biscay just off the northern coast of Spain yielded 2734 ± 2, $2761 + 24/-17$, 1860 ± 4 and $1865 + 23/-25$ Ma U–Pb zircon ages (Guerrot et al., 1989). They are inferred to be part of the Aquitaine block, which underlies most of the Bay of Biscay.

The West African craton is exposed in the Reguibat and Man Shields, which we separated by the Taoudenni Basin. It is made up of two parts: in the west there are Archaean greenstone belts

and granulite–upper amphibolite-facies metamorphic rocks intruded by plutons with wide-ranging compositions; all these rocks being bounded in the east by the early Proterozoic Eburnian Orogen. In the Man Shield, two greenstone belts have been recognized. The oldest (Loko Group) rests upon a granitic basement deformed and metamorphosed during the Leonean Orogeny, dated at $> 2960 \pm 25$ Ma (Pb–Pb whole-rock data; Beckinsale et al., 1980). This was followed by deposition of sediments and volcanics (Kambui Supergroup) which were deformed and metamorphosed during the Liberian Orogeny, dated at between ~ 2850 and 2753 Ma (Rb–Sr whole-rock isochrons: Beckinsale et al., 1980; U–Pb zircon ages: Camil et al., 1983, 1984). Few reliable ages are available from the metamorphic rocks of the Reguibat Shields, but Cahen et al. (1984) have published ages of 320 ± 347 Ma and ~ 2700 Ma based on Rb–Sr whole-rock isochrons. The eastern part of the Man Shield is composed of basement gneisses overlain by metasediments and metavolcanic rocks of the Birrimian Supergroup which was deformed and metamorphosed during the Eburnian Orogeny; all these rocks were in turn unconformably overlain by the Tarkwaian Group molasse. This is separated from the Archaean rocks to the west by a mylonite zone (Sassandra Fault), from which a Rb–Sr whole-rock isochron has given an age of ~ 1670 Ma (Caen-Vachette et al., 1984). Uranium–lead data for pre- to syn-Eburnian granitoids yielded an upper intercept of 2125 ± 25 Ma (Ledent et al., 1969; Legrand, 1971), while late to post-Eburnian granites gave Rb–Sr whole-rock isochron ages of about 2040 ± 45 Ma (Vachette et al., 1975).In the eastern part of the Reguibat Shield, sedimentary and igneous rocks have been deformed and metamorphosed up to amphibolite facies during the Eburnian Orogeny, with an average date of 2022 ± 50 Ma being yielded by Rb–Sr whole-rock isochrons (Cahen et al., 1984). This was followed by intrusion of post-tectonic granites yielding Rb–Sr whole-rock isochron ages of between 1970 ± 46 and 1877 ± 35 Ma (Lassère et al., 1973; Vachette et al., 1973; Sabaté et al., 1977). Inliers with rocks similar to those in the Reguibat Shield also occur to the north within the Anti-Atlas domain, and these have yielded Rb–Sr whole-rock isochrons averaging about 1975 Ma for granitization and migmatization. This was followed by post-tectonic granites ranging in age from 1931 ± 48 to 1692 ± 32 Ma (Charlot, 1976; Cahen et al., 1984), followed in turn by an extensional phase exemplified by formation of the Bou Azzer ophiolites dated at about 672 ± 12 Ma by a Rb–Sr isochron of clay fractions from contact-metamorphosed sediments adjacent to spilitic intrusions comagmatic with the ophiolites (Clauer, 1974). The subsequent phase was one of collisional tectonism which is constrained between the 615 ± 12 Ma age of the deformed Bleida granodiorite and the 578 ± 15 Ma age of the unconformably overlying volcanosedimentary Ouarzazate Group (both U–Pb zircon ages; Ducrot and Lancelot, 1978; Juéry et al., 1974, 1975.)

The eastern margin of the West African craton is bounded by the Trans-Saharan orogenic belt exposed in the Tuareg and Benin–Nigerian Shields (Fig. 6). This orogen formed by opening and closing of an ocean basin between the Tuareg–Nigerian and West African cratons during the late Proterozoic (Black et al., 1979). The opening of the ocean is dated at 785 ± 55 Ma by a Rb–Sr isochron of fine fractions from anchi- and epizonal pelites (Clauer, 1976). Closure lasted from $> 729 \pm 8$ Ma (the U–Pb zircon age of syntectonic calc-alkaline batholiths; Caby and Andreopoulos-Renaud, 1987) until about 604 ± 13 Ma (U–Pb zircon age of the late to post-tectonic Adaf pluton; Bertrand et al., 1978). There is evidence of Archaean events overprinted by Eburnian–Suggarian events in both the Tuareg and Nigerian Shields. Lancelot et al. (1976) therefore analyzed single zircon grains from the In Ouzzal charnockitic paragneiss. The zircons were of two types: brown and rounded, and yellow and euhedral. The brown detrital grains scattered along discordia lines which intersect Concordia at ~ 2050 and 2900–3300 Ma, whereas the yellow grains define a chord with an upper intercept of 2060 Ma. The younger age was interpreted as the age of the granulite-facies metamorphism, while the older age may reflect the age of the source. Evidence of the Trans-Saharan belt in Nigeria is recorded by the 631 ± 21 and 586 ± 5 Ma U–Pb zircon ages on charnockitic granite (Tubosun et al., 1984).

There are several dubious Rb–Sr whole-rock isochrons giving ages between 1175 and 1040 Ma on metamorphic rocks, and these have lead to some suggestions of a tectonothermal event between these ages (Grant et al., 1972; Bertrand and Lassèrre, 1976; Ogezi, 1977).

Geochronological data are scarce in northern South America. However it is sufficient to indicate that the Guyana Shield contains relicts of Archaean rocks remobilized in both the early Proterozoic Trans-Amazonian and late Proterozoic Brasiliano Orogenies (De Almeida et al., 1981; Cordani et al., 1988). The only Archaean ages are: (a) a 2660 ± 30 Ma upper intercept U–Pb from zircons in gneisses the lower intercept is at 545 ± 30 Ma (Gaudette et al., 1978), (b) model Pb–Pb isochrons on whole-rock orthogneisses and feldspars, giving 3770 ± 20 an 3670 ± 10 Ma respectively (Montgomery and Hurley, 1978; Montgomery, 1979), and (c) Rb–Sr whole-rock isochrons yielding ~ 2.7–3.6 Ga (Hurley et al., 1976; Gaudette et al., 1978). However, most of the older dates in the Guyana Shield from both metavolcanics and orthogneisses range between 2.0 and 2.25 Ga: U–Pb zircon ages from gneisses yielded upper intercept ages of 2230 ± 40 and ~ 2020 Ma, while detrital zircons from metagreywackes interbedded with metavolcanics yielded an upper intercept age of ~ 2250 Ma (Gibbs and Olszewski, 1982; Priem, et al., 1978). A Sm–Nd isochron age on metavolcanics gave 2110 ± 90 Ma this being interpreted as the age of extrusion (Gruau et al., 1985); Rb–Sr whole-rock isochrons give a similar range of ages. The other neighbouring shield areas to the south have yielded comparable results, with the additional facet that the rocks of these structures were further remobilized by the late Proterozoic Brasiliano Orogeny between ~ 800 and 600 Ma (e.g. Cordani et al., 1985).

Thus, it is clear that 2.0 and 3.0 Ga source units for the Meguma Group are widespread both in the three cratonic blocks (Baltic, West African and Guyana) as well as in a number of continental blocks within the European Hercynides. The presence of 1.0 Ga detrital zircons in the Cambro-Ordovician metasediments in the Massif Central and Bohemian Massif and the absence of zircons of this age in the Meguma Group suggests that these grains had different source regions and that they cannot be directly correlated. The great extent of the Meguma Terrane suggests that the source area too was large, thus favouring one of the three shield areas as the source area. Of the three shield regions, Baltica would be the least favourable because its southern part is dominated by a ~ 1.0 Ga belt, while central Baltica is dominated by the 1.8–1.88 Ga Svecokarelian Orogen, zircons of neither age group being present as detrital zircons in the Meguma Terrane. Also, late Proterozoic igneous rocks are generally absent in the western Baltic Shield, while detrital zircons of this age are present in the Goldenville Formation. The complete lack of detritus with ages between 1.0 and 2.0 Ga completely excludes a source in eastern North America. Given the present state of knowledge, there does not seem to be any means of further constraining the source of the Goldenville Formation to one or the other of the West African or Guyana Shields. However, the data presented here will become more valuable as new geochronological data are acquired in potential source regions.

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